

1 **Spherulite crystallization induces Fe-redox redistribution in silicic melt**

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15
16 **Abstract**

17
18 Rhyolitic obsidians from Krafla volcano, Iceland, record the interaction between mobile
19 hydrous species liberated during crystal growth and the reduction of ferric iron in the
20 silicate melt. We performed synchrotron μ -FTIR and μ -XANES measurements along a
21 transect extending from a spherulite into optically distinct colorless and brown glass
22 zones. Measurements show that the colorless glass is enriched in OH-groups and
23 depleted in ferric iron, while the brown glass shows the opposite relationship. The color
24 shift between brown and clear glass is sharp, suggesting that the colorless glass zone was
25 produced by a redox front that originated from the spherulite margin and moved through
26 surrounding melt during crystallization. We conclude that the most likely reducing agent
27 is hydrogen, produced by magnetite crystallization within the spherulite. The Krafla flow
28 dramatically captures redox disequilibrium on the microscale and highlights the
29 importance of hydrous fluid liberation and late stage crystallization to the redox signature
30 of glassy lavas.

31
32 *Keywords:* Obsidian, spherulite, oxidation-reduction, FTIR, iron, hydrogen, XANES

33

34 **1. Introduction**

35

36 The chemical and physical processes in magma are influenced by the concentration and
37 reaction of hydrogen-bearing components in the silicate melt. It is well known that the
38 concentrations of molecular H₂O and OH⁻ groups, hereafter referred to collectively as
39 “water”, dissolved in magma govern the positions of mineral liquidus curves and melt
40 viscosities; both rise dramatically, resulting in crystallization (e.g., Geschwind and
41 Rutherford, 1995) and melt stiffening (Hess and Dingwell, 1996), as magma degasses.
42 The relationship between water and the ratio of ferric to ferrous iron in silicate melts
43 remains an active area of research. In H₂-buffered systems, the dissociation of water
44 (e.g., H₂O → H₂ + 1/2O₂, Mueller, 1971) controls oxygen fugacity (*f*O₂); thus *a*H₂O
45 controls Fe³⁺/Fe²⁺. The effect of water as a chemical component at fixed P, T, and *f*O₂ is
46 more nuanced. The study of (Baker and Rutherford, 1996) suggested that the Fe³⁺/Fe²⁺
47 ratio increased with the addition of water; however, as discussed by Wilke et al. (2005)
48 and Gaillard et al. (2001), the study of Baker and Rutherford was not at constant oxygen
49 fugacity (*f*O₂). Experimental studies on rhyolite (Moore et al., 1995) and basalt
50 (Botcharnikov et al., 2005) indicate that water, as a chemical component in a melt held at
51 fixed T, P, and oxygen fugacity (*f*O₂), has no effect on the Fe³⁺/Fe²⁺ ratio. By contrast,
52 other experimental studies indicate that as the bulk water content in increases so too does
53 the Fe³⁺/Fe²⁺ in the melt (Gaillard et al., 2001; Gaillard et al., 2003b). The relationship
54 between water and iron redox state in natural magmas is thus an important factor to
55 consider when reconstructing redox history of degassing magma (e.g., Mathez, 1984;
56 Burgisser and Scaillet, 2007) or deciphering the oxidation states of magma source regions

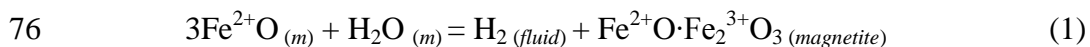
57 (e.g., Carmichael, 1991).

58 Most natural magmas crystallize and vesiculate during their ascent towards and
59 emplacement at the Earth's surface. Because these phase transformations redistribute and
60 ultimately release volatile components from the system, they can drive oxidation-
61 reduction (redox) reactions that will alter the speciation of iron in the magma.

62 Crystallization of silicate minerals that preferentially incorporate ferrous iron in their
63 structures, such as olivine and pyroxene, result in a relatively oxidized melt residuum.
64 The manner by which crystallization of non-ferrous minerals such as quartz and feldspar
65 affects the residual melt redox state is currently unknown; however, because these phases
66 are anhydrous, their growth must redistribute water in the melt (Castro et al., 2008;
67 Watkins et al., 2009), thereby influencing the chemical environment of iron (Gaillard et
68 al., 2003c).

69 The release of H₂ from magma, either through its continuous outward diffusion or
70 by liberation of H₂ gas in bubbles is widely thought to oxidize lavas (e.g., Sato and
71 Wright, 1966; Sato, 1978; Mathez, 1984; Candela, 1986; Christie et al., 1886; Holloway,
72 2004). The diffusive transport of hydrogen out of the melt (*m*) and into bubbles (*v*)
73 displaces the following equilibrium to the right: $2\text{FeO}_{(m)} + \text{H}_2\text{O}_{(m)} = \text{H}_{2(v)} + \text{Fe}_2\text{O}_{3(m)}$.

74 Crystallization of magnetite may similarly alter the oxidation state of basaltic and silicic
75 systems by the "auto-oxidation" reaction as defined by Holloway (2004):



77 The net effect of this reaction on system redox will be determined by the degree to
78 which H₂ gas exits the system. If the hydrogen escapes the magmatic system completely
79 (e.g. to bubbles or the atmosphere), the system is left relatively oxidized. If the hydrogen

80 is retained within the system, however, it may generate relatively reducing fluids adjacent
81 to the zone of crystallization as proposed by Holloway (2004). Here we report adjacent
82 zones of oxidation and reduction in the Krafla flow on a scale of only hundreds of
83 microns.

84 In this paper, we describe fine-scale optical and chemical patterns in obsidian that
85 link spherulitic plagioclase, quartz, and magnetite crystallization to reduction of ferric
86 iron in the rhyolitic melt (Fig. 1). We show that reduced, colorless glass rims jacketing
87 spherulites could have been produced by the expulsion of molecular water from growing
88 spherulites, followed by crystallization of magnetite and concomitant production of
89 hydrogen. The result is a boundary layer of reduced melt that grows with time. Given
90 the considerable sizes (>1 m; Smith et al., 2001) and high volume proportions of
91 spherulites in many rhyolitic lavas and ignimbrites (~ 90 vol.%; Stevenson et al., 1994;
92 Tuffen and Castro, 2009; Tuffen et al., in Review) the interplay of crystallization and
93 micro-scale redistribution of oxidation states has important implications for the evolution
94 of spherulite-rich lavas.

95 **2. Geological Background**

96 Spherulites are radiating, often concentrically arranged crystalline aggregates set
97 in a glassy matrix (Fig. 1). They occur in obsidian domes, vitrophyric ash-flow tuffs
98 (e.g., Smith et al. 2001), large-volume rhyolite flows such as those at Yellowstone (e.g.,
99 Wright, 1915), and in shallow volcanic conduits (e.g., Stasiuk et al. 1996; Tuffen and
100 Castro, 2009). Spherulites nucleate and grow in response to large undercoolings ($>$
101 $200\mu\text{C}$) rapidly imposed on the magma by its degassing and quenching (e.g., Swanson et
102 al., 1989). As dictated by the thermal profile of a magma body (Manley, 1992; Tuffen et

103 al., in Review), spherulitic obsidian develops in spatially restricted zones (e.g., Manley
104 and Fink 1987; Stevenson et al. 1994), comprising a transitional facies that separates the
105 rapidly quenched, outermost vitrophyric rhyolite from a devitrified microcrystalline core.

106 Anomalously high volatile contents exist within and just above the spherulitic
107 zones in lava domes (e.g., Westrich et al. 1988). Castro et al. (2008) have recently shown
108 that the OH⁻ concentrations in glass around spherulites are elevated above the background
109 level. They interpreted these OH⁻ concentration gradients to reflect the combined
110 advection and diffusion of H₂O away from the growing spherulites and numerically
111 modeled these processes in order to estimate crystallization timescales. While their work
112 confirms that spherulite crystallization drives volatile enrichment in silicic glass, it
113 neither identifies the specific form of hydrous species ejected from the growing
114 spherulite, nor constrains why an iron redox shift, manifested as a sharp color difference
115 in the obsidian matrix (Fig. 1), is superimposed on the OH⁻ concentration gradient in the
116 glass. In this paper, we use microscopic chemical and textural evidence collected from
117 the same sample studied by Castro et al. (2008) to demonstrate that regions of elevated
118 OH⁻ are linked to zones of iron reduction in the glass surrounding spherulites. This in
119 turn suggests that spherulite growth may cause changes in the Fe-valence state in rhyolite
120 melt.

121 **3. Analytical and experimental techniques**

122 All analyses were made on a decimeter-sized obsidian sample collected from the
123 Hrafninnuhryggur ridge system on Krafla volcano, Iceland (Tuffen and Castro, 2009).
124 Spherulite mineralogy was determined by 1) microscopic observation, 2) sample
125 magnetism to identify Fe-oxides as magnetite, and 3) compositional data from energy

126 dispersive spectra (EDS) collected on a FEI NOVA nanoSEM600 FEG Variable Pressure
127 Scanning Electron Microscope at the Smithsonian Institution National Museum of
128 Natural History, operated at 7-12 KeV, 5 mm working distance and beam current ranging
129 from 0.5-1 nA.

130 Major element glass compositions were analyzed using a JEOL JXA-8900R
131 electron microprobe (EPMA) running software with ZAF corrections at the Smithsonian
132 National Museum of Natural History. Analyses were performed with an acceleration
133 voltage of 15 keV, a 10 μm beam, and a 10 nA beam current. Standardization was
134 performed on the following natural mineral standards: Quartz (Si), Anorthite (Ca),
135 Bytownite (Al), Microcline (K), Albite (Na), Hornblende (Fe, Mg). A natural rhyolitic
136 glass (VG568, Obsidian Cliffs, Yellowstone, USA) of known major element composition
137 was periodically analyzed to check for instrument drift.

138 Determination of ferrous iron content was via wet chemical analysis of a
139 powdered (grain size~180 μm) obsidian aliquot chipped from the same mass of sample
140 on which all other measurements and observations were made. We attempted to separate
141 out spherulite fragments from the glass powder. However, some glass-encrusted
142 spherulites may have been over looked. Furthermore, it was not possible to separate the
143 thin, clear glass halos from the pervasive brown glass. Thus, these measurements provide
144 an average FeO concentration for the bulk brown glass with very minor dilution by the
145 clear glass fraction. A total of nine (9) analyses were performed following the technique
146 of Peck (1964) with some minor modifications in order to minimize oxidation during
147 sample digestion. Digestion began in 8 ml of fluoroboric acid (HBF_4) for 30 minutes in
148 the ultrasonic bath. We then added 5 ml HF and 2 ml extra of HBF_4 to the solution and

149 completed the digestion under heat for approximately 10 minutes. Three ferrous iron
150 determinations were also made on the U.S.G.S. Glass Mountain Rhyolite standard RGM-
151 1, which yielded a mean value of 1.19 wt. % FeO +/- 0.03 (1 σ). The nominal value for
152 RGM-1 is 1.18 wt. % FeO.

153 H₂O concentrations were determined by synchrotron-FTIR at the Advanced Light
154 Source, Lawrence Berkeley National Laboratory. Measurements were made along
155 traverses oriented perpendicular to the spherulite-glass boundaries on a Thermo Nicolet
156 Magna 760 FTIR spectrometer interfaced with a NicPlan IR microscope (at beamline
157 1.4.3). The IR beam has a diffraction-limited diameter of about 3 μ m. The uncertainty in
158 spot position is \pm 2 μ m. Transmittance spectra were obtained over the mid-IR (1,400-
159 4,000 cm⁻¹) to the near-IR (3,700-6,500 cm⁻¹) regions with MCT detectors, KBr beam-
160 splitters, and the synchrotron light source. 128 scans were used to obtain each spectrum
161 and these spectra were corrected by subtracting a background spectrum collected every
162 hour. We determined OH⁻ concentrations from the intensity of the broad 3,570 cm⁻¹
163 absorption band, utilizing an absorption coefficient of 100 L mol⁻¹cm⁻¹ (Newman et al.
164 1986). We estimate the analytical uncertainty of OH⁻ concentration to be \pm 10% of the
165 measured value.

166 The oxidation state of Fe in the glass was determined at the microscale using Fe
167 K-edge X-ray Absorption Near Edge Structure (μ -XANES) spectroscopy. The area-
168 weighted average energy of the two pre-edge peaks, or centroid, shifts to higher energy as
169 the ratio of ferric to ferrous iron increases, allowing quantification of Fe³⁺/ Σ Fe in silicate
170 glasses (e.g. Berry et al., 2003, Wilke et al., 2005, Cottrell et al., in press). Commensurate
171 with this, the intensity of the peak corresponding to Fe³⁺ (at higher energy) grows

172 proportionately larger relative to the intensity of the peak corresponding to Fe^{2+} (at lower
173 energy); thus the ratio of peak intensities can also be used to quantify oxidation state
174 (Wilke et al., 2005, Cottrell et al., in press).

175 Spectra were collected in fluorescence mode using a 9 element Ge array detector
176 and a silicon channel-cut (311) monochromator at station X26A (bending magnet) at the
177 National Synchrotron Light Source (NSLS), Brookhaven National Lab. The spot size on
178 the sample was $9 \times 5 \mu\text{m}$. Spectra were recorded from 7020-7220eV with a 0.1eV step
179 over the pre-edge from 7106-7118eV at 5s dwell. The pre-edge was deconvolved from
180 the background absorption edge by simultaneously fitting the background with a damped
181 harmonic oscillator function plus a line constrained to have a positive slope and the pre-
182 edge features with two Gaussian peaks. The oxidation state of the glass was quantified
183 using the empirical calibrations in Cottrell et al. (in press) based on pre-edge peak
184 intensity ratios measured on a series of 16 basalt reference glasses with $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios
185 (0.088-0.601) independently determined by Mössbauer spectroscopy and 7 rhyolitic
186 glasses with $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios (0.238-0.806) independently determined by wet chemistry
187 (Moore et al., 1996).

188 We performed a heating experiment on Krafla obsidian in order to induce
189 spherulite crystallization and determine if optical and chemical changes in neighboring
190 melt could result from this crystallization. General experimental details are given here,
191 while more detailed experimental methods are given in the appendix. We first heated a 2
192 cm-edge-length obsidian cube in a split-chamber tube furnace to 770°C at a rate of 65°C
193 min^{-1} . The sample was held at 770°C for 90 minutes. After this period of time, the cube
194 was removed from the furnace and blasted with cold compressed air, which cooled it to

195 room temperature in about 3 minutes. We then extracted a thin (~225 μm), doubly
196 polished wafer from the center of the cube and examined the glass and spherulites near
197 the center of that wafer. We also measured a water concentration profile along a traverse
198 extending from one of the spherulites in the wafer center. We performed a similar
199 experiment at 870°C, but the hydrous melt adjacent to the spherulites vesiculated
200 intensely, rendering comparisons between the experimentally heated and natural
201 unvesiculated samples difficult. Due to a lack of obsidian starting material, we could not
202 repeat the experiment at 770°C.

203

204 **4. Results**

205 *4.1 Analytical measurements*

206 Fig. 1a is a photomicrograph of a natural spherulitic obsidian wafer that is the
207 subject of the measurements discussed in the following paragraphs. The spherulites in
208 this sample are composed of plagioclase and quartz (~95 vol.%), and minor amounts of
209 clinopyroxene (2-3 vol.%) and magnetite (<0.8 vol.%; Castro et al., 2008). All of the
210 spherulites are enclosed in haloes of colorless to light brown glass, which separate them
211 from the pervasive dark brown matrix glass. The transitions between brown and
212 colorless glass tend to be sharp around the larger spherulites (radii>100 μm) whereas the
213 boundaries are more diffuse around the smaller spherulites.

214 Fig. 1b shows a representative OH⁻ concentration profile measured with
215 synchrotron-FTIR by Castro et al. (2008). The area under the profile is proportional to
216 the amount of OH⁻ groups in the silicate glass surrounding the spherulite. Castro et al.
217 showed that the OH⁻ concentration increases with the spherulite size, and typically

218 matches the amount of water that would be ejected during complete crystallization of
219 anhydrous minerals from a melt volume equal to that of the spherulite. Differences
220 between the measured and predicted OH⁻ contents show that some spherulites retained
221 water during their growth, consistent with the presence of glass and microvesicles (Castro
222 et al., 2008).

223 Table 1 shows representative major element glass compositions measured within
224 and across the differently colored glass regions. These data show that the composition of
225 glass in the colorless halo, the transitional zone, and the distal brown glass surrounding
226 the spherulite in Fig. 1a are indistinguishable within analytical error. Most importantly,
227 there is no discernable change in the total iron content as the spherulite-glass boundary is
228 approached; i.e., glass color is not a function of the total iron content but rather the ratio
229 of ferric to ferrous iron. Fig. 2 is a graph showing the major element compositions of
230 glass along a line traverse emanating from the spherulite shown in Fig. 1a.

231 By analogy with glasses equilibrated experimentally under controlled fO_2 , the
232 color difference between the halo and distal glass in the natural sample was suspected to
233 reflect the proportions of ferric and ferrous iron, with the darker brown glass having more
234 ferric iron than the light-colored glass (Gaillard et al. 2002; Donald et al., 2006;
235 Moriizumi et al., 2008). This hypothesis was confirmed with μ -XANES measurements.
236 Spectra were collected on the spherulite shown in figure 1a both within the clear glass
237 rim, approximately 20 μ m from the spherulite-glass border, and in the far-field brown
238 matrix glass where the OH⁻ concentration flattens out, about 330 μ m from the spherulite
239 edge. Figure 3ab shows the raw pre-edge spectra, model components, and total model
240 fits to the spectra. The spectrum taken in the distal brown glass (spot 2) displays a

241 proportionately larger Fe^{3+} peak, indicative of a greater contribution from ferric iron. This
242 can be seen even more clearly in figure 3c when the baseline subtracted spectra are
243 superimposed. Quantitatively, the centroid (area-weighted average of the two pre-edge
244 peaks) shifts to higher energy by 0.18 eV moving from the clear rim to the distal brown
245 glass. Consistent with this, the ratio of pre-edge peak intensities (i.e. $I(\text{Fe}^{3+})/I(\text{Fe}^{2+})$
246 $I(\text{Fe}^{3+})/[I(\text{Fe}^{2+}) + I(\text{Fe}^{3+})]$) shift to reflect a greater contribution from ferric iron in the
247 distal brown glass than in the halo, consistent with a relative change in the ratio of
248 $\text{Fe}^{3+}/\sum\text{Fe}$ of 0.06. The μXANES results clearly indicate a difference in the relative Fe
249 oxidation state of the two glass regions, with the clear glass halos demonstrably reduced
250 relative to the brown matrix glass. The absolute ferric iron content determined for the
251 distal brown glass ($\text{Fe}^{3+}/\sum\text{Fe} = 0.23\pm 0.04$) compares favorably with, but is more reduced
252 than, the wet chemical determination of $\text{Fe}^{3+}/\sum\text{Fe} = 0.29\pm 0.02$, whereas Fe in the clear
253 brown glass is more reduced ($\text{Fe}^{3+}/\sum\text{Fe} = 0.17\pm 0.04$).

254 Fig. 4 shows measurements of the widths of several colorless glass haloes versus
255 their corresponding spherulite radii. Estimated uncertainties in rim widths, stemming
256 from the diffuse nature of the glass color boundaries, are shown as vertical error bars.
257 The nonlinear regression to the data is a power-law function relating rim width to
258 spherulite radius. Also shown in Fig. 4 are the predicted reduction rim widths for each
259 spherulite based on calculations using empirical relationship of Gaillard et al. (2003a),
260 which relates reduction front position or width to hydrogen fugacity, temperature (800°C)
261 and time. These model calculations will be discussed further below.

262

263 *4.2 Experimental results*

264 The optical character and OH⁻ concentration of the heated obsidian are shown in
265 Figs. 5 and 6. To aid comparisons with the natural state, we also include images of
266 unheated spherulite-rim combinations collected on the spherulite shown in Fig.1. The
267 transmitted-light photomicrographs were taken at the same illumination and focal depth,
268 and the imaged wafers were about the same thickness.

269 The salient features and changes in the heated sample are: 1) thin, ~10 μm wide
270 veneers of plagioclase (identified with EDS on an SEM) jacketed all spherulites in the
271 sample, 2) thin brown fringes composed of glass, plagioclase, and magnetite crystals
272 located just inboard of the plagioclase veneers, 3) the colorless glass rims brightened, and
273 on their outer periphery, thin, light-brown glass zones developed that appear to have
274 propagated toward the spherulites, 4) circumferential cracks formed parallel to the
275 colorless-brown glass boundaries, in some cases directly on them, and 5) along many of
276 these cracks, small, ~10-30 μm, “bubble-trains” grew, comprising groups of spherical
277 (Fig. 5c) to nearly completely flattened vesicles (Fig. 5d).

278 The most apparent changes that occurred during the heating experiment were
279 crystallization and brightening of the colorless haloes (Fig. 5a, b). That the
280 crystallization was “new” is indicated by the fact that none of the unheated spherulites
281 have a jacket of feldspar around them. The crystallization rate implied by the amount of
282 new plagioclase ($\sim 1.7 \times 10^{-9}$ m sec⁻¹; ie., width of plagioclase veneer divided by the
283 experiment time) is similar to the average value ($\sim 1.9 \times 10^{-9}$ m sec⁻¹) determined by Castro
284 et al. (2008) from their diffusion models at 800°C.

285 The circumferential fractures and small vesicle trains resulted from processes
286 occurring, respectively, below and above the rheological glass transition (T_g). At the

287 given heating rate, the sample reached T_g ($\sim 700^\circ\text{C}$; Castro et al., 2008) in ~ 10 minutes.
288 During this short time the glass may have fractured due to its greater thermal expansion
289 compared to the mineral phases in the spherulite. If this were the case, then fractures
290 formed in tension as the expanding glass pulled away from the spherulite. Once the
291 fractures formed, and the glass passed through T_g , bubble nucleation occurred
292 preferentially along the fractures, which channeled volatile components to the growing
293 bubbles. The cracks apparently served as conduits for degassing of the volatile-enriched
294 boundary layers. An FTIR profile (Fig. 6) measured across one of these fractures
295 indicates that the dissolved OH^- content decreases in the vicinity of a fracture, suggesting
296 degassing into that fracture.

297

298 **5. Mechanism of Fe-reduction during crystallization**

299 It is clear from the geometry of the colorless glass haloes, specifically their
300 proportional increase in width with spherulite size (Fig. 4), and their mimicking of the
301 spherulite shapes, that the colorless glass rims resulted from the growth of the spherulites.
302 The μ -XANES spectra indicate that some (~ 5 -7% absolute or 26% relative) of the ferric
303 iron in the glass adjacent to the spherulites has been reduced to ferrous iron (Fig. 3). In
304 this section we address how these redox changes could have occurred in light of the
305 analytical and experimental evidence.

306 We first consider the possibility that the heterogeneous ferric-ferrous iron
307 distribution in the glass is related to the cooling history. Metrich et al. (2006) showed
308 that the ferric-ferrous ratio in peralkaline rhyolitic glass inclusions appeared to increase
309 upon slow cooling over several hundred $^\circ\text{C}$ in high temperature μ -XANES experiments.

310 They attributed the apparent oxidation in XANES spectra upon cooling to a change in the
311 coordination environment of iron; high temperature favors tetrahedrally coordinated
312 ferrous iron while lower temperatures stabilize octahedrally coordinated ferric iron.
313 Metrich et al conclude that the XANES spectra of slowly cooled samples will appear
314 more oxidized due to this coordination shift. To account for the reduced halos in the
315 spherulitic obsidian, the reduced halos would have had to cool much more rapidly than
316 the more oxidized distal brown glass, which is physically unlikely because latent heat
317 liberated from the growing spherulites would have slowed cooling of the melt adjacent to
318 the spherulite (Tuffen et al., submitted). Moreover, the color change in the glass supports
319 a real difference in the oxidation state of iron, not a coordination change. For these
320 reasons, this mechanism was probably not important for the formation of the colorless
321 rims.

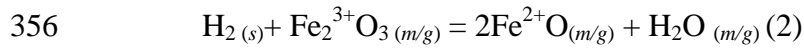
322 We next consider the possibility that the reduced-Fe signature in the rims arose
323 from the late-stage crystallization of magnetite. The distribution of magnetite within
324 spherulites, mainly as radial aggregates sandwiched between larger domains of
325 plagioclase and quartz (Fig.1), suggests that magnetite grains formed in the latest stage of
326 crystallization, from an interstitial melt that would have been enriched in water, further
327 stabilizing magnetite (Sisson and Grove, 1993). Magnetite crystallization could act to
328 reduce the melt adjacent to the spherulite via two potential mechanisms. First, hydrogen
329 produced by magnetite crystallization in the presence of water (i.e., “auto oxidation,”
330 reaction 1) would necessarily diffuse out of the spherulite, and would have subsequently
331 reduced ferric iron in the neighboring melt. Second, magnetite crystallization alone will
332 reduce iron in the glass residuum by virtue of its higher ferric/ferrous ratio (just as olivine

333 crystallization would raise the ferric/ferrous ratio). We show below that either
334 mechanism, or both in concert, could have generated the reduced halos.

335 “Auto-oxidation” as described by Holloway (2004) may proceed in hydrous
336 silicates when magnetite with a higher ferric/ferrous ratio than the melt from which it is
337 crystallizing becomes stable. Pure end-member magnetite has a $\text{Fe}^{3+}/\Sigma\text{Fe}$ of about 0.67,
338 compared to the value of 0.22 - 0.29 in the brown glass; therefore magnetite precipitation
339 could have resulted in “auto oxidation” in this hydrous lava from Krafla. In this scenario,
340 magnetite precipitation proceeds through the consumption of water, generating one mole
341 of H_2 for every mole of magnetite crystallized (reaction 1). If H_2 leaves the system
342 completely through degassing along fractures, the remaining material is left relatively
343 oxidized. As H_2 diffuses through a melt, however, it necessarily results in a reduction
344 front. We believe that these spherulites capture this disequilibrium state.

345 Simple mass balance arguments confirm the plausibility of this scenario. The
346 modal proportion of magnetite in the spherulites, determined by BSE image analysis on
347 10 different spherulites, is about 0.3 vol.% (± 0.12). The amount of Fe^{3+} now residing in
348 the magnetite grains within the large spherulite pictured in Fig. 1a is about 0.003 mg Fe^{3+}
349 (5.38×10^{-8} moles of Fe^{3+}), and because these magnetite grains contain some Ti (Castro
350 et al., 2008), this estimate of Fe^{3+} in magnetite is a maximum. If all of the Fe^{3+} residing
351 in the magnetite were created by oxidation of Fe^{2+} in the liquid, magnetite crystallization
352 would produce a maximum of 2.7×10^{-8} moles of H_2 via reaction (1). This hydrogen could,
353 in turn, reduce 5.38×10^{-8} moles of ferric iron (i.e. 0.003 mg) in the adjacent melt (and
354 produce water) according to:

355



357

358 Subscripts above refer to: (s)-spherulite and (m/g)-melt/glass. If all of the H₂
359 were transferred to the 100µm halo jacketing the spherulite in Figure 1a, it would shift
360 the percentage of Fe³⁺ from ~22% (measured via XANES) to 9%. We observe Fe³⁺ in the
361 halo equal to 16% (table 1), indicating that “auto-oxidation” within the spherulite could
362 actually be responsible for iron reduction in the halo. This calculation is a maximum
363 because it assumes that 100% of the ferric iron in the magnetite had to be converted from
364 FeO (i.e. reaction 1). We observe, however, that the brown glass contains about 22%
365 ferric iron. Assuming that reaction (1) proceeded at 78% efficiency (i.e. 22% of the ferric
366 iron incorporated into the magnetite was already ferric), we still find that the percentage
367 of Fe³⁺ would drop to 11%. Within the error of the XANES measurements, the
368 uncertainty of the reaction efficiency and mass balance, and even taking into account that
369 the magnetite is Ti-bearing, this scenario remains plausible.

370 The auto-oxidation scenario is also corroborated by the high concentration of
371 water in the halos. We observe the number of moles of “H₂-equivalent” currently residing
372 in the halo (i.e. one half the moles of OH⁻, calculated from the OH⁻ concentration of
373 0.19% measured by FTIR) equal to 5x10⁻⁸. This is ~45% higher than the number of
374 moles of H₂ expected from magnetite crystallization via reaction (1). We therefore
375 observe more “water” (hydroxyl and molecular) in the glass surrounding these spherulites
376 than produced by magnetite crystallization. This “excess” water is expected because
377 water is also rejected as an incompatible element in the growing spherulite (Castro et al.,
378 2008).

379 A second mass-balance argument can be made that magnetite precipitation alone
380 (i.e. not relying on the presence of water) could have caused the reduction halos.
381 Magnetite incorporates iron in a ratio of $2\text{Fe}^{3+}:1\text{Fe}^{2+}$. Only 0.4% magnetite
382 crystallization (as a percent of the glass now occupied by the reduced halos) is required to
383 cause the ferric iron to decrease from ~22 to the observed value of ~16%. If we assume
384 that all the magnetite seen in the spherulites crystallized from the volume of glass now
385 occupied by the reduced halos, we observe 0.7% crystallization (which would reduce the
386 ferric iron to 11%). If we assume that all of the magnetite seen in the spherulites
387 crystallized from the volume of glass that now comprises the volume occupied by the
388 spherulites plus the reduced halos, we observe 0.3% crystallization (which would reduce
389 the ferric iron to 18%). These two values therefore bracket the maximum and minimum
390 extents of magnetite crystallization, and either extreme provides a reasonable mechanism
391 by which the glass in the halos could have been reduced. Crystallization of 0.4%
392 magnetite from the halo would simultaneously cause the total iron concentration (FeO) in
393 the halo to fall from 3.2 to 2.8%. This is close to within the error of the microprobe
394 measurements. We have the added uncertainty of not knowing if the iron concentration
395 could have been increased in the halo volume due to its incompatibility in early-
396 crystallizing quartz and plagioclase. Nevertheless, no decrease or gradient in the iron
397 concentration is observed in the halos relative to the distal brown glass. This, combined
398 with the demonstrable presence of water in the system and the likelihood of the auto-
399 oxidation reaction proceeding, leads us to slightly favor the auto-oxidation scenario as a
400 means of generating the halos, but neither can be ruled out and in fact both could have
401 contributed.

402 Gaillard et al. (2003a, c) simulated the process of hydrogen-flux iron reduction by
403 exposing natural Fe-bearing rhyolitic melt and glass cylinders to reducing atmospheres of
404 hydrogen and hydrogen-argon gas mixtures. Their experiments produced many of the
405 features we observe around the natural spherulites: 1) a sharp change in glass color
406 bounding a zone of reduced $\text{Fe}^{3+}/\text{Fe}_{\text{total}}$ in hydrous glass, 2) sigmoidal OH^- concentration
407 profiles emanating from the sample edge and attributable to hydrogen incorporation in
408 the melt followed by diffusion of molecular water along the concentration gradient, and
409 3) an offset between the reaction front position and the point of elevated OH^- in the glass.

410 Gaillard et al. (2003a, c) proposed that because hydrogen is very reactive with
411 iron in the silicate melt, the reaction front progress in the melt is governed by the
412 solubility and diffusion of hydrogen in the melt. The effective diffusion rate of H_2 , in
413 turn, was limited by the $f\text{H}_2$ of their experiments. Gaillard et al. observed that the rate of
414 progress of the reduction front ($\mu\text{m}'\text{s}/\text{hour}$) was several orders of magnitude slower than
415 the expected hydrogen diffusivity ($\mu\text{m}'\text{s}/\text{sec}$) in the melt.

416 Gaillard et al. (2003a) performed time series experiments in order to characterize
417 the rate of advancement of the reduction front with time. They observed a square-root-
418 of-time dependence of the front position and, based on linear relationships between the
419 square of the reduction front position and run duration, they extracted reduction rate
420 constants, $K=\mu^2/t$, where μ is the reaction front position, and t is time, for their
421 experiments at 800°C and a range of hydrogen fugacities (0.02-50 bars). They proposed
422 that the reduction rate was limited by hydrogen incorporation, which in turn, is a function
423 of the fugacity-dependent H_2 solubility and diffusivity in the melt.

424 It is possible that the spherulites behaved in a similar manner, that is, they acted

425 as hydrogen point sources to the neighboring melt. Although the boundary conditions are
426 slightly different, e.g., the spherulites are an internal, as opposed to external hydrogen
427 source, the empirical kinetic data of Gaillard et al. (2003c) can be used to assess whether
428 the widths of the natural reduction rims are compatible with the timescales of spherulite
429 growth. In other words, we assume that the reduction front started moving at the onset of
430 spherulite growth and stopped when the growth ceased. By this simple scenario, the rim
431 widths depend on the growth timescale, and therefore the duration that the expelled
432 hydrogen had to react with the rhyolite melt according to the reaction rate constant K
433 (Gaillard et al. 2003a).

434 Castro et al. (2008) estimated average spherulite growth rates by modeling the
435 combined advection and diffusion of water away from the growing spherulites and fitting
436 model profiles to the natural OH^- concentration profiles. Even though their model did not
437 account for hydrogen incorporation in the melt and its unknown effect on the kinetics of
438 H_2O diffusion, their growth rates match those determined experimentally on
439 compositionally similar melts (e.g., Baker and Freda, 2001), and are probably accurate to
440 an order of magnitude. According to their results at 800°C the spherulites grew at an
441 average rate of $\sim 10^{-9}\text{m/s}$ (Castro et al., 2008), which is similar to the rate determined in
442 our spherulite growth experiments at 770°C (appendix).

443 Results of rim-width calculations are shown in Fig. 4. Individual spherulite
444 growth timescales were determined by dividing the average growth rate (800°C ; Castro et
445 al., 2008) by their spherulite radii. Using the individual growth timescales (t), we
446 calculated rim widths (μ) using the relation $\mu = (K \cdot t)^{1/2}$. The best match between the
447 calculated and natural rim widths was attained with the K value derived from the lowest

448 H₂ fugacity ($f_{\text{H}_2}=0.02$ bar) in the experiments of Gaillard et al. (2003a). We found a very
449 large mismatch at their next highest f_{H_2} (~0.25 bar H₂) and attribute this to the nearly one
450 order of magnitude increase in K at $f_{\text{H}_2}=0.25$ bar, which translates to comparatively rapid
451 rim growth relative to the spherulite growth timescales.

452 The agreement between the calculated and real Fe-reduction rim widths is good,
453 suggesting that the natural rims could have developed under conditions of relatively low
454 hydrogen fugacity (~0.02 bar according to Gaillard et al., 2003a). This match supports
455 our hypothesis that it is molecular hydrogen that extrudes from the growing spherulite,
456 thereby fluxing the ferric iron in the neighboring melt and causing the propagation of a
457 redox front.

458

459 **6. Experimentally induced spherulite growth and Fe-reduction**

460 The experimental results of Gaillard et al (2003a, c) show that hydrogen fluxing
461 of a silicate melt or glass will cause reduction of the ferric iron and the formation of a
462 sharp reduction front that progresses with the square root of time. The close agreement
463 between these experimental results and our natural observations suggests that the
464 colorless haloes formed as a result of ferric iron reduction during spherulite growth, with
465 hydrogen being a permissible reductant.

466

467 **7. Conclusion**

468 Our analysis indicates that the spherulitic growth of anhydrous phases can cause
469 significant changes in Fe-oxidation state in the neighboring melt or glass around the
470 growing crystals. Fe-redox reactions are driven by the liberation of hydrogen, and we

471 propose that it is hydrogen produced as a product of magnetite crystallization. The result
472 is an Fe-reduction reaction that propagates through the glass or melt with time. That
473 spherulite crystallization can cause reduction of ferric iron in the silicate melt suggests
474 that this phenomenon could prevent oxidation of silicic lavas during their emplacement
475 despite extensive crystallization. For example, Carmichael (1991) indicates that
476 voluminous post-caldera rhyolite lavas at Yellowstone National Park, USA underwent
477 little to no change in redox state compared to earlier-erupted ash flows. Carmichael
478 (1991) used the apparent lack of oxidation of these lavas to support his hypothesis that
479 “silicic magmas have redox states that reflect their source regions rather than H₂ loss.”
480 He further explained that the low bulk H₂O content of the post caldera lavas could have
481 suppressed the activity of the H₂O as most of the water would be speciated as OH groups
482 (Stolper, 1982).

483 Spherulite crystallization is widespread in the interiors of the Yellowstone
484 rhyolites (Wright, 1915; Colony and Howard, 1934), with some flows containing well
485 over 50 vol.% spherulites (e.g., in the Nez Pierce flow). We suggest that the lack of
486 variation in redox state may have been enabled by spherulite crystallization and local
487 hydrogen solute rejection, but retention of generated hydrogen within the flows. Thus
488 Carmichael’s (1991) conclusion that “silicic magmas with small amounts of iron and
489 large amounts of water do not have their redox states reset” upon eruption may be a
490 consequence of the offsetting effects of crystallization and glass reduction at the
491 microscale.

492

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499

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624

625 **Figure Captions**

626

627 **Figure 1.** A) Photomicrograph of spherulites (round and elliptical, black) in obsidian
628 (brown). Various analytical points and traverses are shown (emp=electron probe
629 microanalyzer; FTIR=Fourier Transform Infrared Spectroscopy; XANES=x-ray
630 absorption near-edge spectroscopy). Black diamonds indicate approximate locations of
631 μ -XANES measurements. Broader dark swaths are tracks left by a laser ablation ICPMS
632 (these data are not discussed here). B) A back-scattered electron image showing typical
633 internal spherulite texture. Bright phases are clinopyroxene and magnetite. C) An OH⁻
634 concentration profile determined by FTIR collected along the lower right traverse in
635 frame A. The vertical grey line is the approximate location of the colorless-brown glass
636 boundary.

637

638 **Figure 2.** Major element compositions of glass leading up to a spherulite margin (at
639 position = 0 μ m). All analyses were collected with an electron probe microanalyzer. The
640 vertical grey bars demarcate the approximate (\pm 10 μ m) position of the colorless-brown
641 glass boundary.

642

643 **Figure 3.** Raw μ -XANES spectra of the pre-edge region, model components, and model

644 fits for two points within the glassy matrix depicted in Figure 1A (a) Point 1 inside the
645 colorless glass halo and (b) point 2 in the distal brown glass. (c) Model fits of the spectra
646 in (a) and (b) baseline-subtracted and superimposed. Approximate positions of analysis
647 points are shown on the inset photomicrograph. These data show the relative changes in
648 ratio of Fe^{2+} to Fe^{3+} in the colorless and brown matrix glass. Iron in the glass near the
649 spherulite (point 1) is significantly reduced compared to that in the far-field brown glass
650 (point 2).

651

652 **Figure 4.** Colorless-glass rim width versus the apparent spherulite radius, measured on
653 photomicrographs of the Krafla obsidian (solid dots). Error in rim width, shown as
654 vertical bars, ranges from about 10-20%, and originates from the diffuse nature of the
655 colorless-brown glass boundaries. Regression curve is a power law function of the form:
656 $y=2x^{0.69}$; $R^2=0.97$. Solid triangles are the predicted rim widths based on the
657 experimentally constrained hydrogen-iron reaction rate constant of Gaillard et al.
658 (2003b), and the assumption that the rims developed during the interval of time that the
659 spherulites grew, which in turn, is taken from the growth rate data of Castro et al. (2008).

660

661 **Figure 5.** Photomicrographs (500X; 200 μm f.o.v.) collected on: A) a natural spherulite-
662 glass combination, and B-D) an experimentally heated spherulitic obsidian. Frame A)
663 shows the natural appearance of the spherulite margin (at top) and colorless and brown
664 glass as viewed in transmitted, plane polarized light. The wafer is 197 μm thick. Frame
665 B) shows the typical appearance of the spherulite and colorless and brown glass after
666 heating a similar piece of obsidian to 770 μC for approximately 90 minutes. This wafer is
667 about 202 μm thick. Note the thin veneer ($\sim 10 \mu\text{m}$) of plagioclase coating the spherulite
668 margin (upper right) and the relative brightening of the colorless glass halo. Note also
669 the crack running roughly in the center of the colorless halo with a string of small
670 vesicles on its tip. Frame C) shows another view of the incipient crystallization and
671 vesiculation that occurred during heat treatment. The vesicles here formed along an
672 arcuate fracture in the colorless glass zone. Frame D) is a backscattered electron image
673 of the three zones developed at the spherulite margin: the brown microlite rich fringe (a),
674 plagioclase veneer (b), and the colorless and Fe-reduced matrix glass (c).

675

676 **Figure 6.** An OH^- concentration profile measured along a traverse extending from a
677 spherulite in the heated obsidian sample. The position of traverse is shown on the
678 subjacent photomicrograph. The vertical dashed and solid grey lines on the profile bound
679 the light brown glass zone that may have formed as a result of degassing of hydrogen into
680 the crack shown in the photomicrograph. Note the coincidence between the depression in
681 the OH^- concentration profile and the position of the crack at $\sim 75\text{-}80 \mu\text{m}$.

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